# Global Change and the Dark of the Moon

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June 1992

JSR-91-315



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| **REPORT DOCUMENTATION PAGE** | *Form Approved* *OMB No. 0704-0188* |
|---|---|

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| 1. AGENCY USE ONLY (*Leave blank*) | 2. REPORT DATE | 3. REPORT TYPE AND DATES COVERED |
|---|---|---|
| | June 30, 1992 | Final |

**4. TITLE AND SUBTITLE**

Global Change and the Dark of the Moon

**5. FUNDING NUMBERS**

PR - 8503Z

**6. AUTHOR(S)**

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**7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES)**

The MITRE Corporation
JASON Program Office A10
7525 Colshire Drive
McLean, VA 22102

**8. PERFORMING ORGANIZATION REPORT NUMBER**

JSR-91-315

**9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES)**

Department of Energy
Office of Program Analysis
Office of Energy Research
Washington, DC 20585

**10. SPONSORING/MONITORING AGENCY REPORT NUMBER**

JSR-91-315

**11. SUPPLEMENTARY NOTES**

**12a. DISTRIBUTION/AVAILABILITY STATEMENT**

Distribution unlimited; open for public release.

**12b. DISTRIBUTION CODE**

**13. ABSTRACT** (*Maximum 200 words*)

We have considered the possibility of using earthshine to measure the reflectance properties of the earth (albedo and phase function). Measurements of earthshine carried out by Danjon in 1926--33 show that even then the average albedo could be determined with a precision of ± 0.01 and that both synoptic and seasonal variations could be observed clearly.

We show that, after correction for wavelength dependence and the opposition effect in the lunar reflectance properties, Danjon's visual albedo of 0.40 can be reconciled with the ERBE satellite Bond albedo of 0.30. We recommend a modern earthshine monitoring program (advantages include global integration, continuous coverage, ground basing, and low cost) as a complement to present and planned satellite measurements.

**14. SUBJECT TERMS**

earth albedo, earthshine, elementary photometry, danjon's measurements

**15. NUMBER OF PAGES**

**16. PRICE CODE**

| 17. SECURITY CLASSIFICATION OF REPORT | 18. SECURITY CLASSIFICATION OF THIS PAGE | 19. SECURITY CLASSIFICATION OF ABSTRACT | 20. LIMITATION OF ABSTRACT |
|---|---|---|---|
| UNCLASSIFIED | UNCLASSIFIED | UNCLASSIFIED | SAR |

NSN 7540-01-280-5500 Standard Form 298 (Rev 2-89)
Prescribed by ANSI Std Z39-18

# Contents

# 1 EXECUTIVE SUMMARY

We have considered the possibility of using earthshine to measure the reflectance properties of the earth (albedo and phase function). Measurements of earthshine carried out by Danjon in 1926–33 show that even then the average albedo could be determined with a precision of $\pm$ 0.01 and that both synoptic and seasonal variations could be observed clearly.

We show that, after correction for wavelength dependence and the opposition effect in the lunar reflectance properties, Danjon's visual albedo of 0.40 can be reconciled with the ERBE satellite Bond albedo of 0.30. We recommend a modern earthshine monitoring program (advantages include global integration, continuous coverage, ground basing, and low cost) as a complement to present and planned satellite measurements.

# 2 INTRODUCTION

In this report we consider a "new" monitor of global change: observations of earthshine as a measure of the earth's albedo and phase function. This is actually the oldest method of determining the earth's albedo [1], and was studied extensively from 1926–33 by Danjon [2, 3]. Very little has been done since [4], we suspect in part because of lack of intense interest in the subject and because of the advent of satellite measurements. However, earthshine has several clear advantages in a modern context, as we hope to demonstrate.

We begin our presentation with a brief discussion of the importance of the earth's albedo to climate. We then turn to a qualitative discussion of earthshine and review the relevant notions of photometry. This is followed by a review of Danjon's measurements, both to illustrate the method and to show what could be done even with 1920s technology. We note several corrections to Danjon's observations and show how they can be reconciled with satellite measurements. Finally, we discuss some possibilities for modern earthshine measurements.

# 3 THE EARTH'S ALBEDO AND ITS CLIMATE

The earth's climate system is basically a large heat engine. Energy comes into the system in the form of short-wavelength radiation from the sun, peaking at a wavelength of 0.5$\mu$m (a black-body temperature of about 6000 K) (see Figure 1). Almost 99 percent of the sun's radiation is contained in the so-called short wavelength region of 0.15 to 4.0 $\mu$m. Of this energy 46 percent is in the infrared region above 0.74 $\mu$m, 9 percent in the ultraviolet below 0.4 $\mu$m and the remaining 45 percent in the visible, 0.4 to 0.74 $\mu$m. A significant fraction of this energy is absorbed by the earth, where it drives the motion of the atmosphere and oceans before being radiated back into space as long-wavelength radiation peaking at a wavelength of 15 $\mu$m (a black-body temperature of $T_E = 255$ K).

The power going in to the earth's climate system is

$$P_{in} = C\pi R_E^2(1 - A) \ , \tag{3-1}$$

where $C = 1370$ W/m$^2$ is the solar constant, $R_E = 6378$ km is the earth's radius, and $A = 0.30$ is the earth's shortwave (Bond) albedo, giving the fraction of the incident shortwave solar radiation (between 0.15 and 4.0 $\mu$m) that is reflected from the earth without being absorbed. For this latter quantity, we have adopted the value determined by ERBE satellite measurements [6]. Similarly, the longwave power that the planet radiates into space is

$$P_{out} = 4\pi R_E^2 \sigma\epsilon T_E^4 \ , \tag{3-2}$$

where $\sigma$ is the Stefan-Boltzmann constant and $\epsilon$ is the emissivity at the top of the atmosphere (about 5.5 km, where the long wave radiation is emitted).

If the planet is in radiative equilibrium, these two powers are equal,

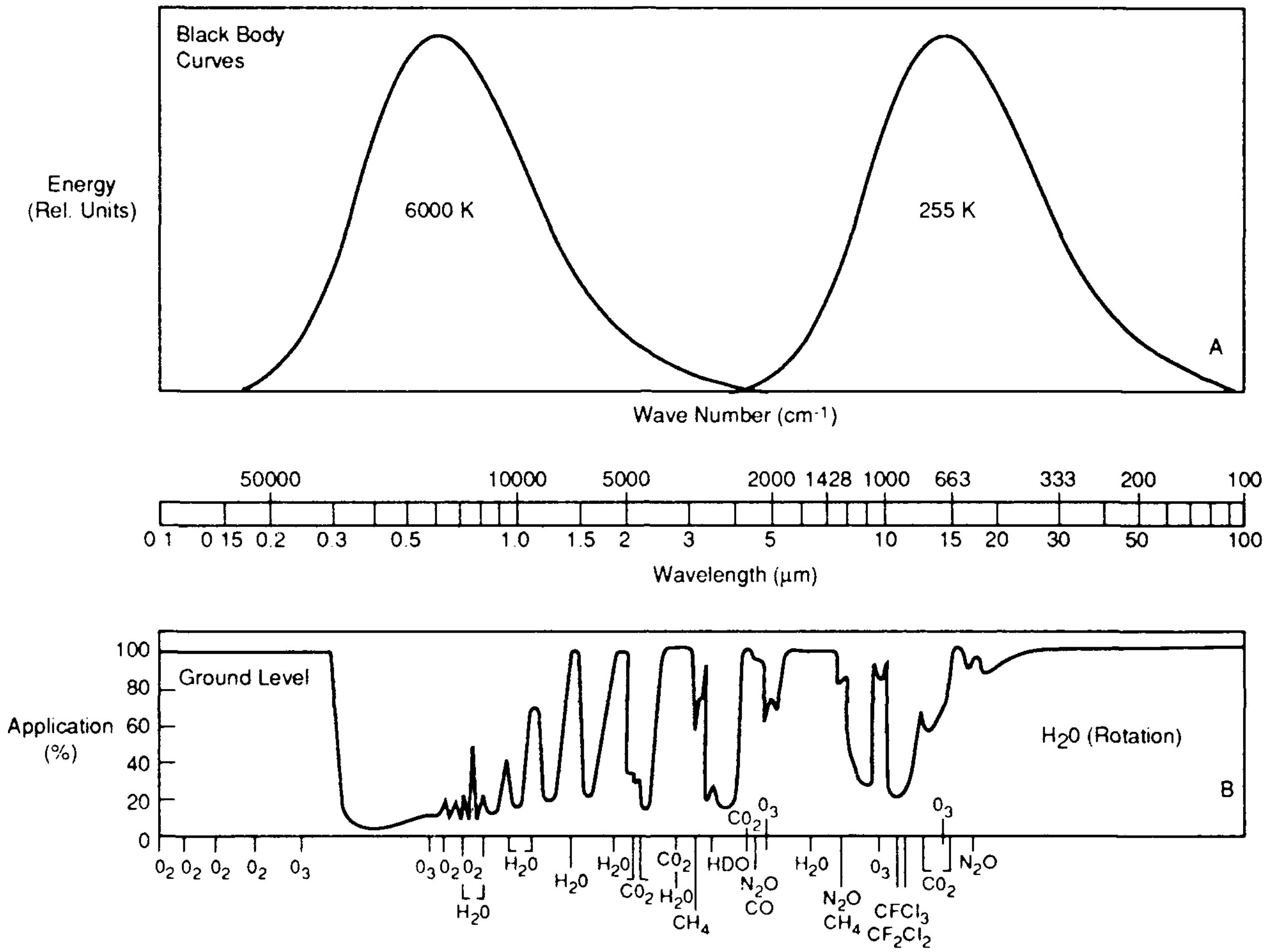



**Figure 1.** (A) Spectral distribution of longwave emission from black bodies at 6000 K and 255 K, corresponding to the mean emitting temperatures of the sun and earth, respectively, and (B) percentage of atmospheric absorption for radiation passing from the top of the atmosphere to the surface. Notice the comparatively weak absorption of the solar spectrum and the region of weak absorption from 8 to 12 μm in the longwave spectrum (from Reference [5]).

giving

$$T_E^4 = \frac{C}{4\sigma\epsilon}(1 - A) . \tag{3–3}$$

Thus, the global shortwave albedo directly controls the earth's temperature.

Most (about 70 percent) of the solar energy entering the climate system is absorbed by the surface or within the atmosphere, with the balance being reflected. The energy budgets of the three principal kinds of surfaces that cover the earth differ dramatically. The heat capacity of the oceans is very large and the solar radiant energy can penetrate efficiently. The oceans thus respond slowly to changes in solar flux and act as the regulators of the climate system. Snow and ice reflect a large fraction of the incoming radiation (albedo up to 0.8). The albedo does not change until near the melting point, when the optical character of the surface begins to alter. Because of the phase change, snow and ice have a large effective heat capacity and influence primarily the slow physics of the atmosphere. Land surfaces respond most rapidly to changes in solar radiation. Their effective heat capacity is low because visible light does not penetrate and thermal conduction is very slow. The albedo of the land surface depends on the angle of the incident radiation and, when the land is covered with vegetation, on the spectrum of the incident radiation.

The global albedo depends upon the reflectance properties of each surface element of the earth. Any good quality radiation detector (e.g., pyranometer) can be used to measure albedo. Two instruments, one looking upward and the other downward, will provide an instantaneous measure of downwelling and upwelling radiation. The instruments may be at ground level or on an aircraft or satellite, depending on the area to be viewed.

The value obtained by an instrument will depend on the zenith angle of the sun, the transmittance of the atmosphere, the nature of the surface, and the nature of the cloud cover, if any, including the thickness of the cloud, the water content of the cloud, and the droplet size distribution within the

cloud. The angle of the reflecting surface to the horizontal, particularly if it is facing towards or away from the sun, is important. The surface albedo is high for dry, light-colored, smooth surfaces and is low for wet, dark-colored, rough surfaces. In the case of vegetation the surface albedo will depend on the height of plants, percentage of ground cover, angle of leaves and the leaf area index. Figure 2 shows that, on average, the air and land each contribute about 25 percent of the reflectivity, with clouds making up the other 50 percent.

Typical albedos for clear land and ocean are 0.16 and 0.08, respectively, while the corresponding overcast values are 0.50 and 0.44; the albedo for clear desert is 0.23, while that for snow is 0.68. Local albedos show considerable synoptic and diurnal variability (see Figure 3). The global albedo also shows a seasonal variability, in part because of the greater land area in the Northern Hemisphere (see Figure 4).

Although it is the (effective) top-of-the-atmosphere temperature that appears in Equation (3-3), the surface temperature should also vary with albedo. This is borne out by the data presented in Figure 5, where the monthly mean global surface temperatures are plotted against the monthly mean global albedo for 100 years of an atmospheric general circulation model (GCM) run with the sea surface temperatures having a fixed seasonal cycle [7]. Perhaps coincidentally, the linearization of Equation (3-3) ($\Delta A/\Delta T_E = -0.01\ \mathrm{K}^{-1}$) is a reasonable description of the data. A similar conclusion can be drawn from the observations shown in Figure 6, where the monthly mean global temperatures for 1985 are plotted against the monthly mean global albedos determined by ERBE [6].

Greenhouse warming scenarios give ambiguous predictions of the likely change in the albedo. On the one hand, the increasing water vapor in the atmosphere could increase the cloudiness and hence the albedo. However,

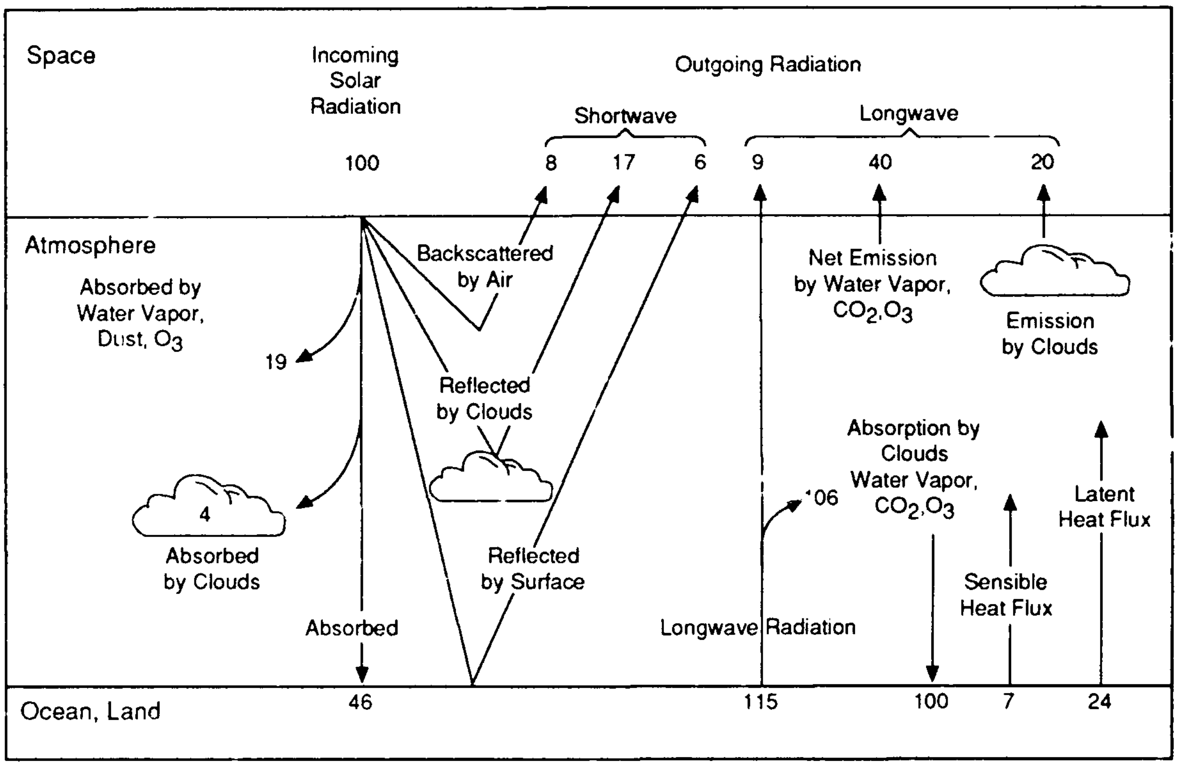



**Figure 2.** Schematic representation of the atmospheric heat balance. The units are percent of incoming solar radiation. The solar fluxes are shown on the left-hand side, and the longwave (thermal IR) fluxes are on the right-hand side (from Reference [5]).

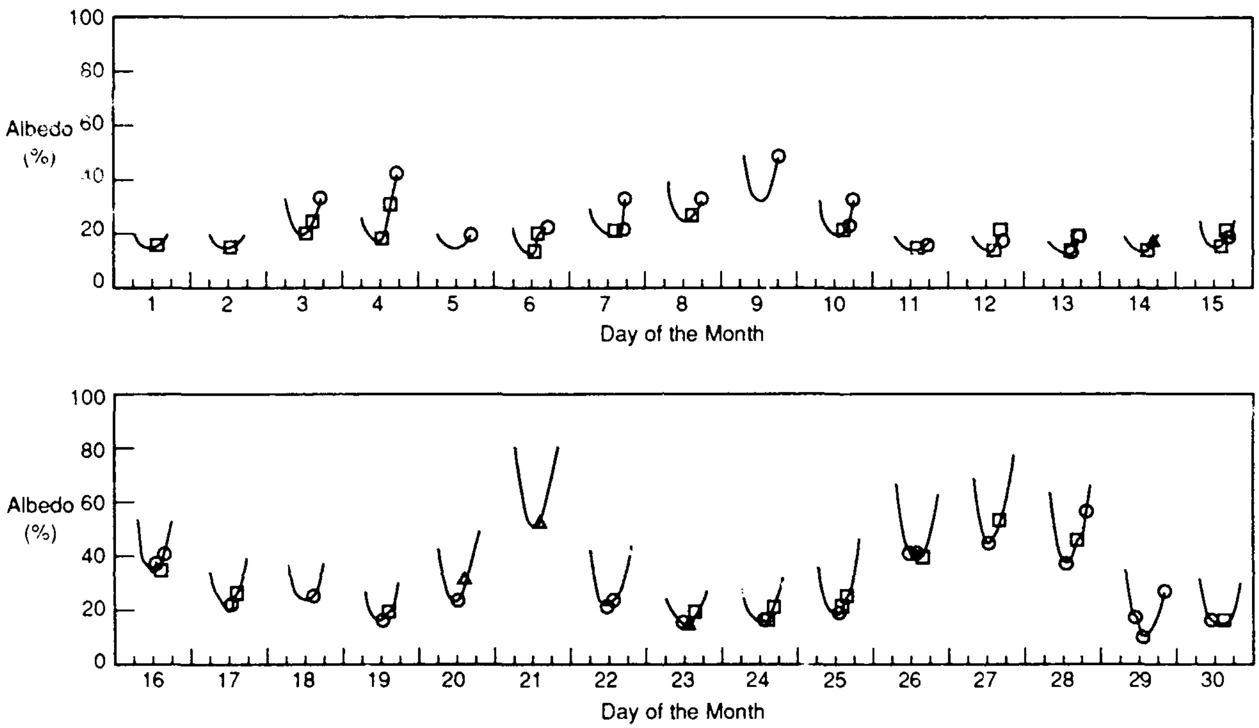



**Figure 3.** Time history of April 1985 albedo for a region in the western Pacific Ocean (long. = 148.75°W, lat. = 21.25°S) (from Reference [6]).

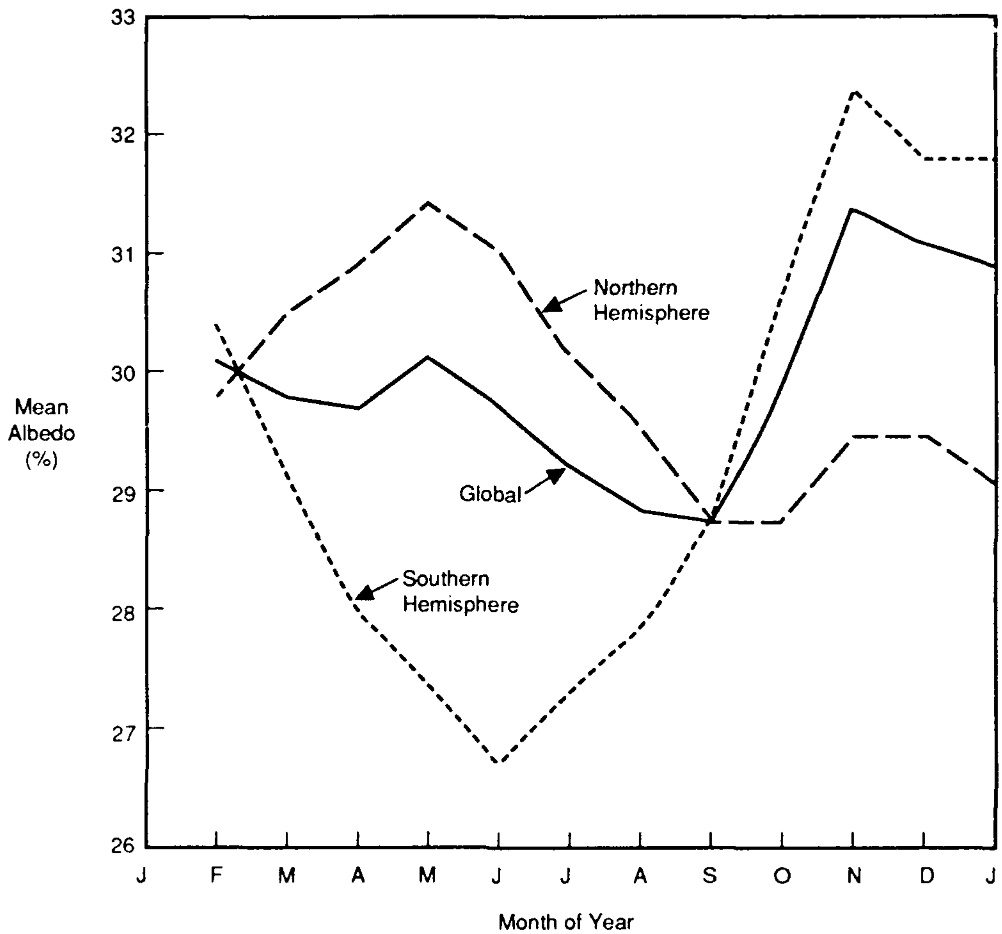



**Figure 4.** Annual variation of ERBE monthly mean albedo for hemispheres and the globe (from Reference [6]).

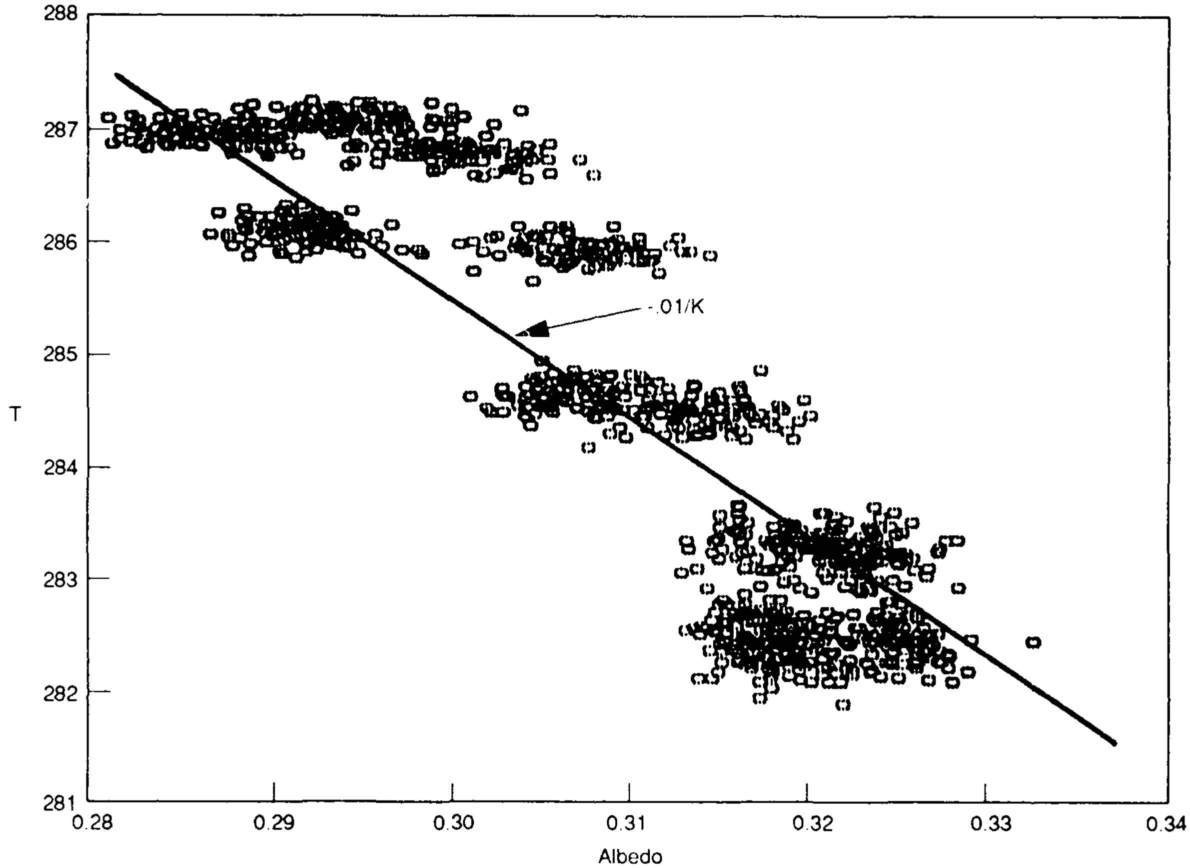



**Figure 5.** Correlation of the monthly mean globally averaged surface temperature with the monthly mean global albedo in a 100-year run of the NCAR CCM-1 with climatological sea-surface temperatures (from Reference [7]).

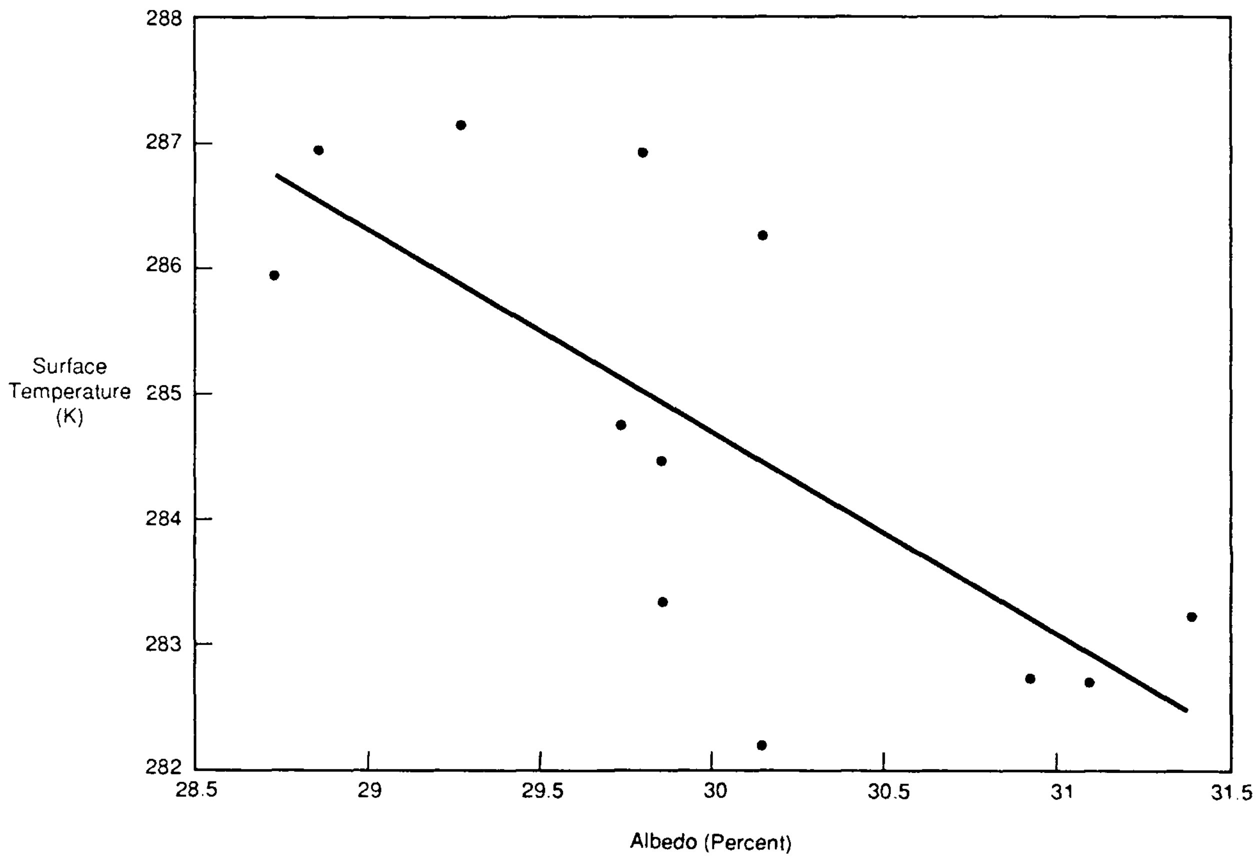



**Figure 6.** Correlation of the monthly mean global surface temperature with the ERBE monthly mean albedo for 1985. The least-squares fit line shown has a slope of -0.016/K.

the decreasing snow and ice coverage will act to decrease the albedo. The net effect need not be negative (as might be expected from Equation (3-3) as the temperature increases), as the emissivity of the atmosphere will change with increasing greenhouse gas content.

The most precise measurements of the earth's albedo come from satellite measurements such as those of ERBE [6, 8]. Here an instrument measures the amount of outgoing shortwave radiation for one spot on the earth at one particular solar zenith angle from a given viewing elevation and azimuth. Complex "scene" models are then used to convert this measurement into a total flux of outgoing shortwave radiation (i.e., that going into all viewing directions) and hence an albedo; further modeling is used to average over the diurnal cycle (zenith angle dependence). Finally, all pixel values are averaged to obtain a global value. The process is quite complex, with many modeling assumptions involved. Other drawbacks of such measurements include the expense and risk of satellites and the difficulty of maintaining a very accurate calibration (better than 1 percent) in a space-based instrument. Although great effort has been expended to ensure the accuracy of satellite-determined albedos, an independent check would be, at the least, reassuring.

# 4 EARTHSHINE

Earthshine is sunlight that is reflected by the earth to the moon (see Figure 7). It therefore contributes to the illumination of the moon beyond that of the much more intense direct sunshine and is most easily visible as a ghostly glow of the dark portion of the lunar disk. The phenomenon was known to the ancients and understood by Kepler in 1604.

The geometry of the sun-earth-moon system is most simply characterized by the sun-moon-earth angle $\iota$, the phase angle of the moon (see Figure 8). Because the earth-moon distance is much smaller than the earth-sun distance (ratio $\approx 2.5 \times 10^{-3}$), the sun-earth-moon angle, or phase angle of the earth, is $\phi \approx \pi - \iota$. It is clear that earthshine will be most easily visible when $\iota$ is largest (crescent moon as seen on earth and full earth as seen on the moon), although groundbased measurements for the very largest $\iota$ are precluded by daylight. Conversely, earthshine is most difficult to observe when the moon is nearly full and the earth is a crescent ($\iota \approx 0$).

The use of a coronagraph enables observation of the earthshine as close as 38 hours before full moon [9], even though at this time the earthshine comes from a very narrow crescent of the earth and scattering and diffraction effects are relatively strong. The ratio of earthshine to sunshine visible from the earth thus varies throughout the lunar cycle; it is less than about $10^{-3}$ under the most favorable conditions.

To first order, earthshine observations measure scattered light in the plane of the ecliptic. The assumption of incident azimuth independence makes the observations general. The relations between the longitudes of the sun, the moon, and the observation station are described in Figures 9 and 10.

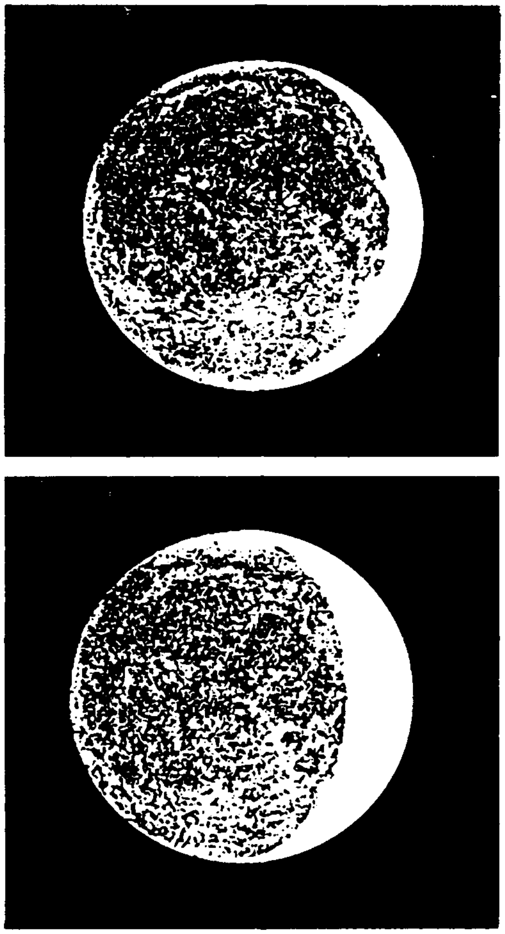


**Figure 7.** Photographs of the earthshine at various lunar phases (from Reference [2]).

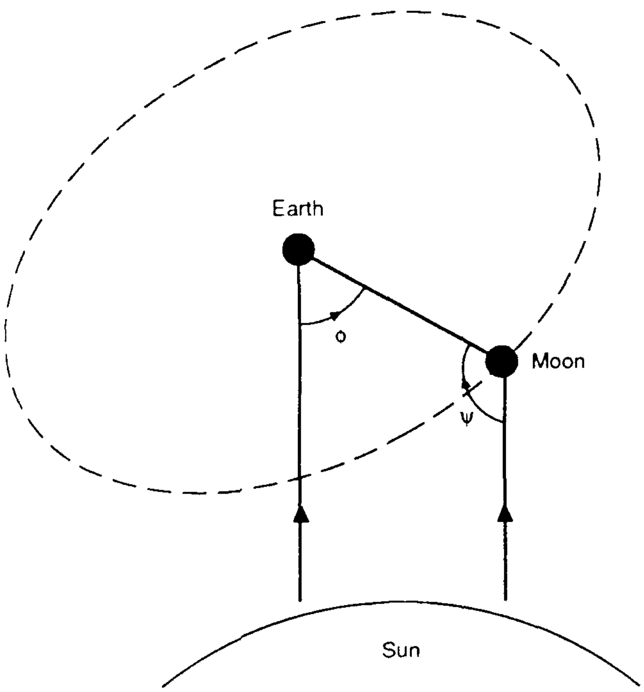



**Figure 8**. Geometry of the earth-moon-sun system. The earth's phase angle $\Phi$ is related to the moon's phase angle $\Psi$ as $\Phi \approx \pi - \Psi$.

Earthshine clearly depends upon the visible reflectance properties of the earth and thus can be used to determine the earth's albedo. Such measurements were pursued by Very in 1912 [1] and more extensively by Danjon [2]. In a modern context they offer a number of attractive advantages, including an instantaneous coverage of a large region of the globe, the potential for nearly continuous long time series of observations, and the fact that they are ground based (and are hence relatively inexpensive and easily maintained, calibrated, and upgraded).

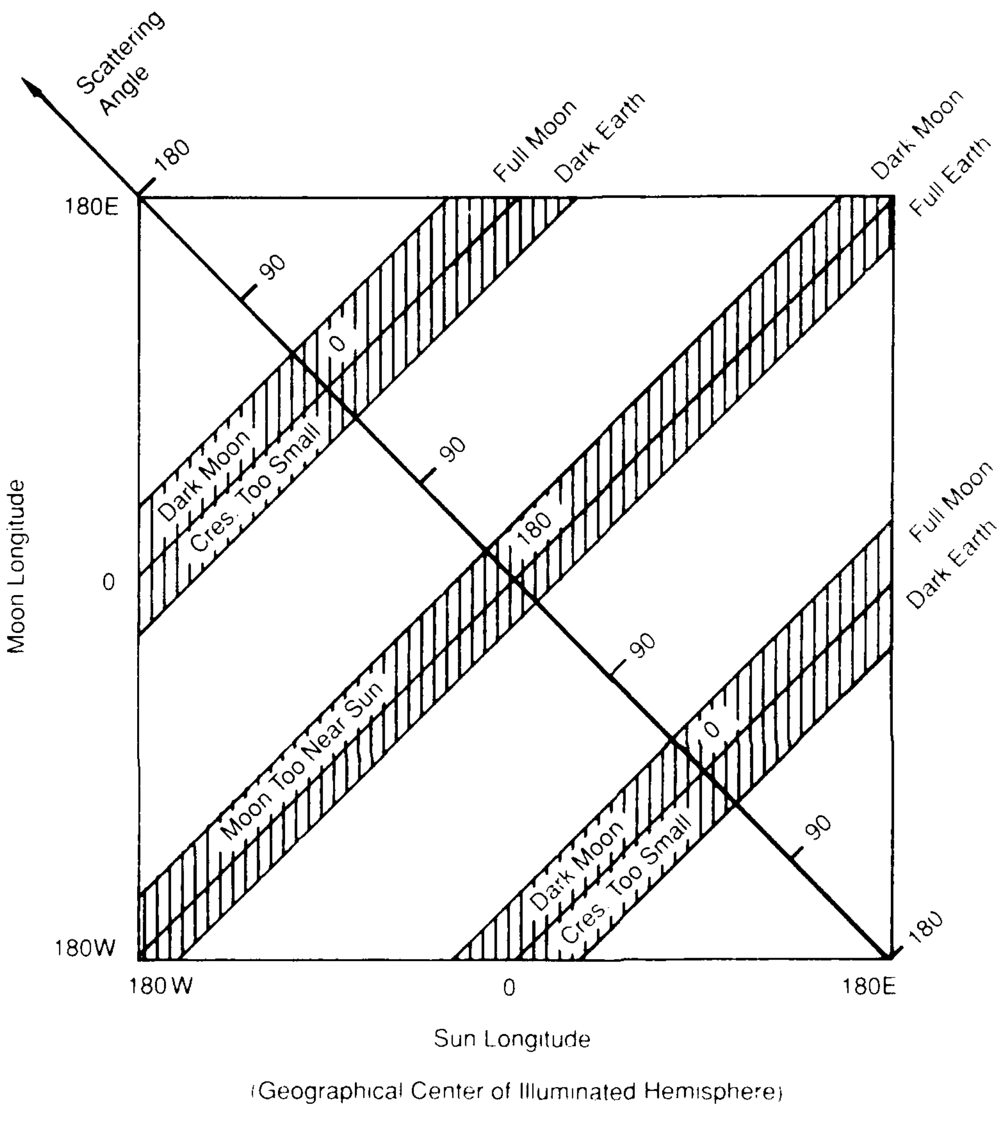



**Figure 9.** Geometrical relationships involved in earthshine measurements. The tilt of the earth's axis has been ignored. The earth's albedo is a function of the Sun longitude (which part of the earth is being illuminated) and time. The albedo is proportional to the integral of the scattered light over all directions; earthshine observations give measurements of the angular distribution in the plane of the ecliptic. The angle of scattering is indicated along the diagonal. Measurements are difficult or impossible too near either a full moon or a new moon, indicated by hatched areas.

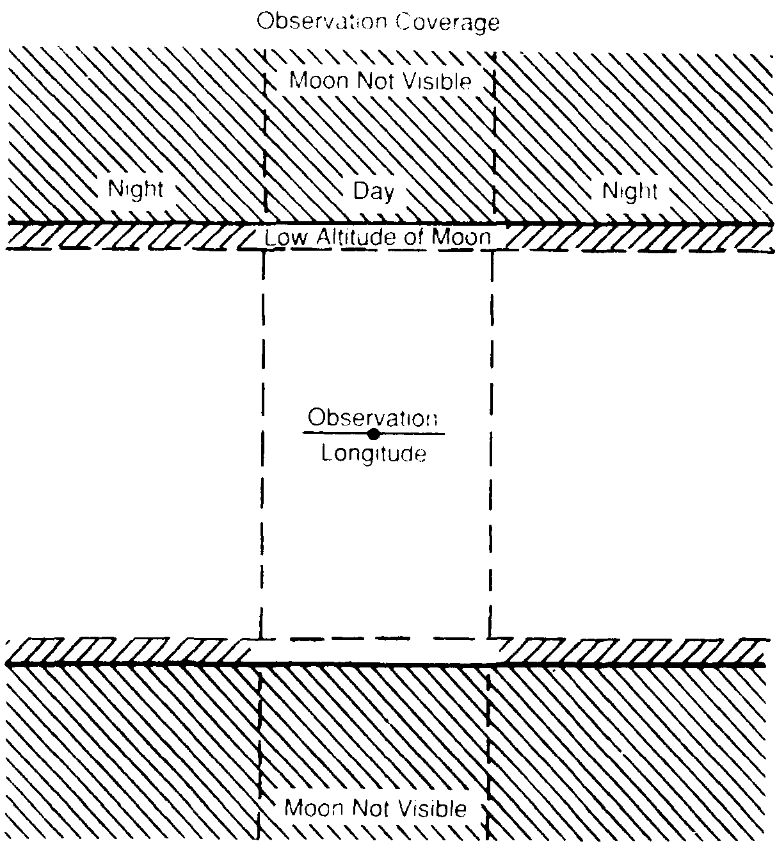



**Figure 10.** Coverage of the Sun/Moon longitude diagram (Figure 9) possible from a given observation point near the equator. Figure 10 is meant to be overlaid on Figure 9 with the Observation Longitude point in the middle of Figure 10 placed on the positive diagonal of Figure 9 at the point where both the Sun and the Moon longitudes are equal to the chosen Observation Longitude. Then earthshine observations are possible in the clear areas; in the hatched areas of Figure 10 the moon is either below the horizon (and hence not visible) or is so close to the horizon that the earth's atmosphere interferes with the measurements. The regions of day or night are indicated. Some measurement systems (but not all) will be usable in the daytime.

# 5 ELEMENTARY PHOTOMETRY

In order to describe the use of earthshine to determine the earth's albedo, it is necessary to review some elementary notions of photometry. Consider, as shown in Figure 11, a plane surface of area $S$ illuminated uniformly by light making an angle $\theta_i$ with the normal and observed at a similarly defined angle $\theta_r$ and azimuth $\chi_r$. In general, the cross section for the plane to effect this scattering will depend upon all three angles, plus the azimuth of the incoming light. Dependence on the latter is usually ignored.

However, for a perfectly diffuse scatter (Lambert surface), the cross section is given by the azimuth-independent expression

$$\frac{d\sigma}{d\Omega_r} = Sr\frac{1}{\pi}\cos\theta_i\cos\theta_r \ , \qquad (5-1)$$

where $\Omega_r$ is the solid angle of scattering and $r$ is the reflectance of the surface material. As expected, the total cross section for scattering is $\int d\Omega_r(d\sigma/d\Omega_r) = Sr\cos\theta_i$, the product of the reflectance and the projection of the illuminated area on the direction of illumination. We also note that when $\theta_r = \theta_i$, $d\sigma/d\Omega_r = (Sr/\pi)\cos^2\theta_i$.

Now consider the more complex situation of the earth illuminated by the sun. For a given phase angle $\phi$, the cross section for light scattering can be written as

$$\frac{d\sigma}{d\Omega} = \pi {R_E}^2\frac{1}{\pi}pf_E(\phi) \ . \qquad (5-2)$$

Here, the earth's phase function $f_E$ is defined so that $f_E(0) = 1$, implying that the cross section for the earth to backscatter the sunlight (and hence the intensity of the full earth) is given by $p{R_E}^2$. The quantity $p$ is called the geometric albedo and depends upon the reflectance properties of

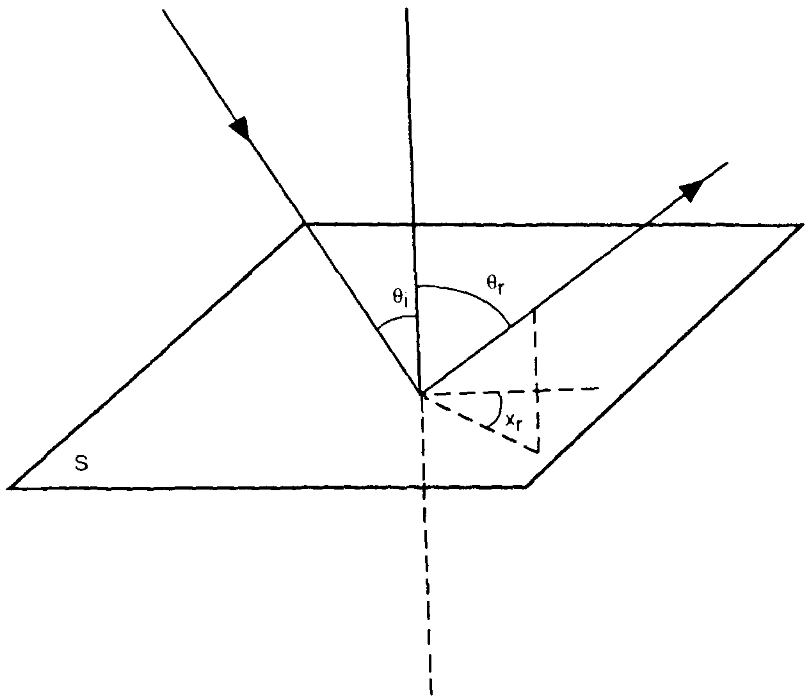



Figure 11. A plane reflector of area S illuminated at an angle $\theta_i$ and observed at angle $\theta_r$ with azimuth $x_r$.

the earth's surface and the geometry of a sphere. From its definition, it can be seen that $p$ is the ratio of the light backscattered by the sphere to that backscattered by a normally illuminated perfectly reflecting ($r = 1$) Lambert disk of the same area ($\pi R_E^2$). For a Lambert sphere, $p = 2r/3$ and $f(\phi) = (\sin\phi + (\pi - \phi)\cos\phi)/\pi$. It should be noted that since the earth has a variegated surface, both $p$ and $f_E$ will depend upon the particular hemisphere illuminated.

The Bond albedo is, by definition, the ratio of the total cross section to the area of the planet's disk ($\pi R_E^2$). Thus,

$$A = pq\ ; \qquad q = \int_{-\pi}^{\pi} f_E(\phi)\sin\phi d\phi\ , \tag{5-3}$$

where $q$ is termed the phase integral. A basic difficulty in albedo measurements is thus apparent: any one observation of the reflected light (whether a given view from a satellite or the earthshine at a given phase of the moon) determines the cross section at only one scattering angle, so that some extrapolation to all scattering angles must be performed to obtain the entire phase function and hence the phase integral. In satellite measurements, this is done by scene models; for the earthshine, it is done partially by measuring $f_E$ at various phases. More generally, for earthshine it is necessary to make assumptions about (or models for) the sunlight scattered out of the ecliptic.

Many decades of photometry have determined the phase function, geometric albedo, phase integral, and visual albedo for various solar system objects. For example, the phase function of the moon as determined by Rougier [10] is shown in Figure 12, taken from [11].

Selected values of the phase integral and albedo are: Mercury (0.563,0.055), Venus (1.296,0.61), earth (1.095,0.40), and the moon (0.584,0.073), as given by Danjon [3].

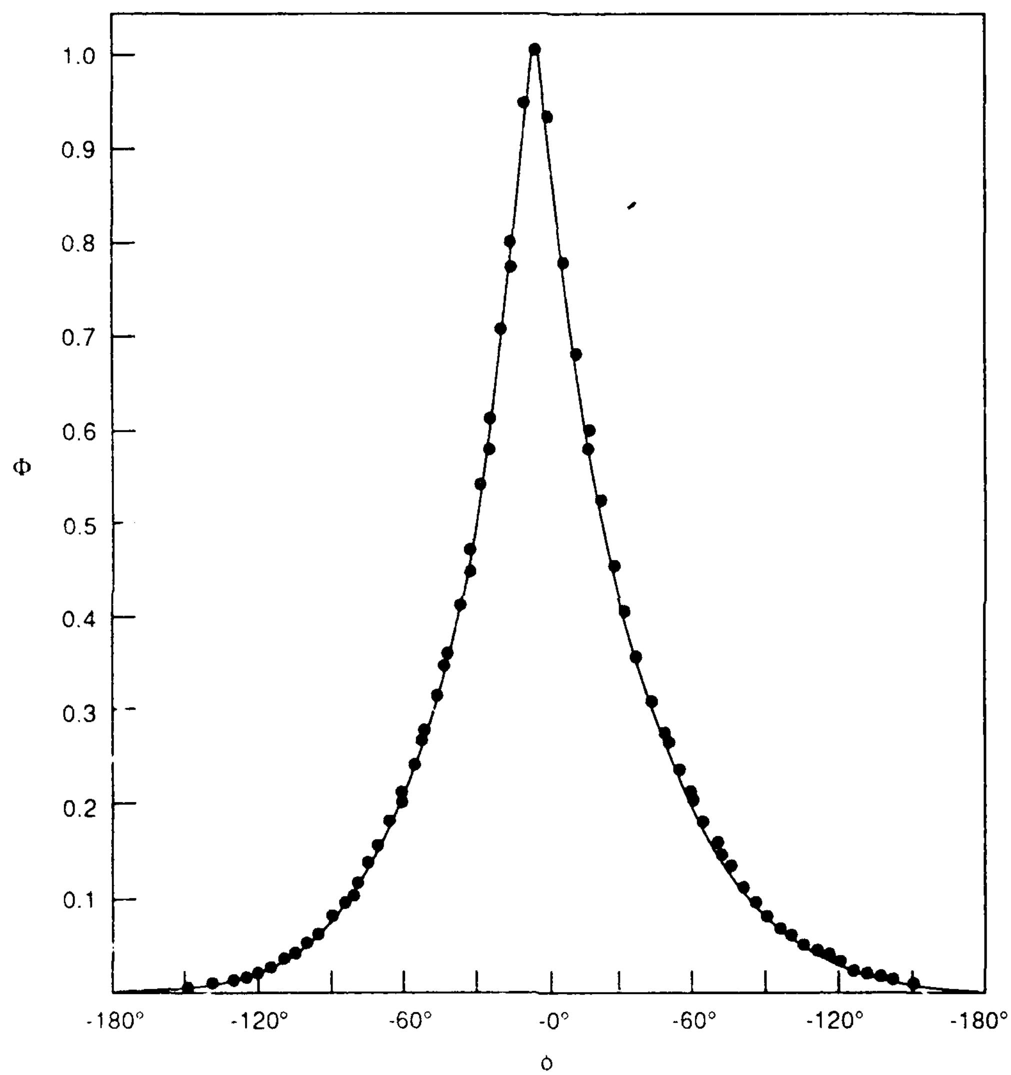



Figure 12. Integral phase function of the moon (from Reference[10]).

# 6 DANJON'S MEASUREMENTS

Danjon's measurements of earthshine [2, 3] involved comparing the intensities of two well defined patches of the lunar surface (denoted by $A$ and $B$), with one in the sunshine and the other in the earthshine. The patches were chosen to be bright (highland) with similar optical properties and almost diametrically opposite near the lunar limb (see Figure 13). A photometer produced adjacent images of the moon, so that patch $A$ in the earthshine was adjacent to patch $B$ in the sunshine (see Figure 14). An adjustable diaphragm ("cat's eye") reduced the intensity of the light from $B$ to allow the ratio of the intensities, $D_{AB} = I_A/I_B$, to be measured for various phase angles. For opposite phase, the roles of the patches were reversed.

Such differential measurements removed many of the uncertainties associated with atmospheric absorption, varying solar constant, etc.

Since the earthshine is backscattered from patch $A$ while the sunshine scatters from $B$ with phase angle $\psi$, the observed ratio of the intensities is

$$D_{AB} = \frac{I_E \frac{1}{2} p_A f_A(0)}{I_S \frac{1}{2} p_B f_B(\psi)} = \frac{I_E}{I_S} \frac{1}{f_M(\psi)} \qquad (6\text{-}1)$$

where $I_E$ and $I_S$ are the intensities of the earth and sun as observed on the moon, $p_{(A,B)}$ are the geometric albedos, and $f_{(A,B)}$ are the phase functions. The second equality follows from assuming that $p_A = p_B$ and that $f_A = f_B = f_M$, a common phase function for both lunar patches. Note that if these assumptions were not valid, there would be systematic differences between observations during the positive and negative phases of the moon.

Solving Equation (6-1) for the ratio of intensities yields

$$\frac{I_E}{I_S} = D_{AB} f_M(\psi) \, . \qquad (6\text{-}2)$$

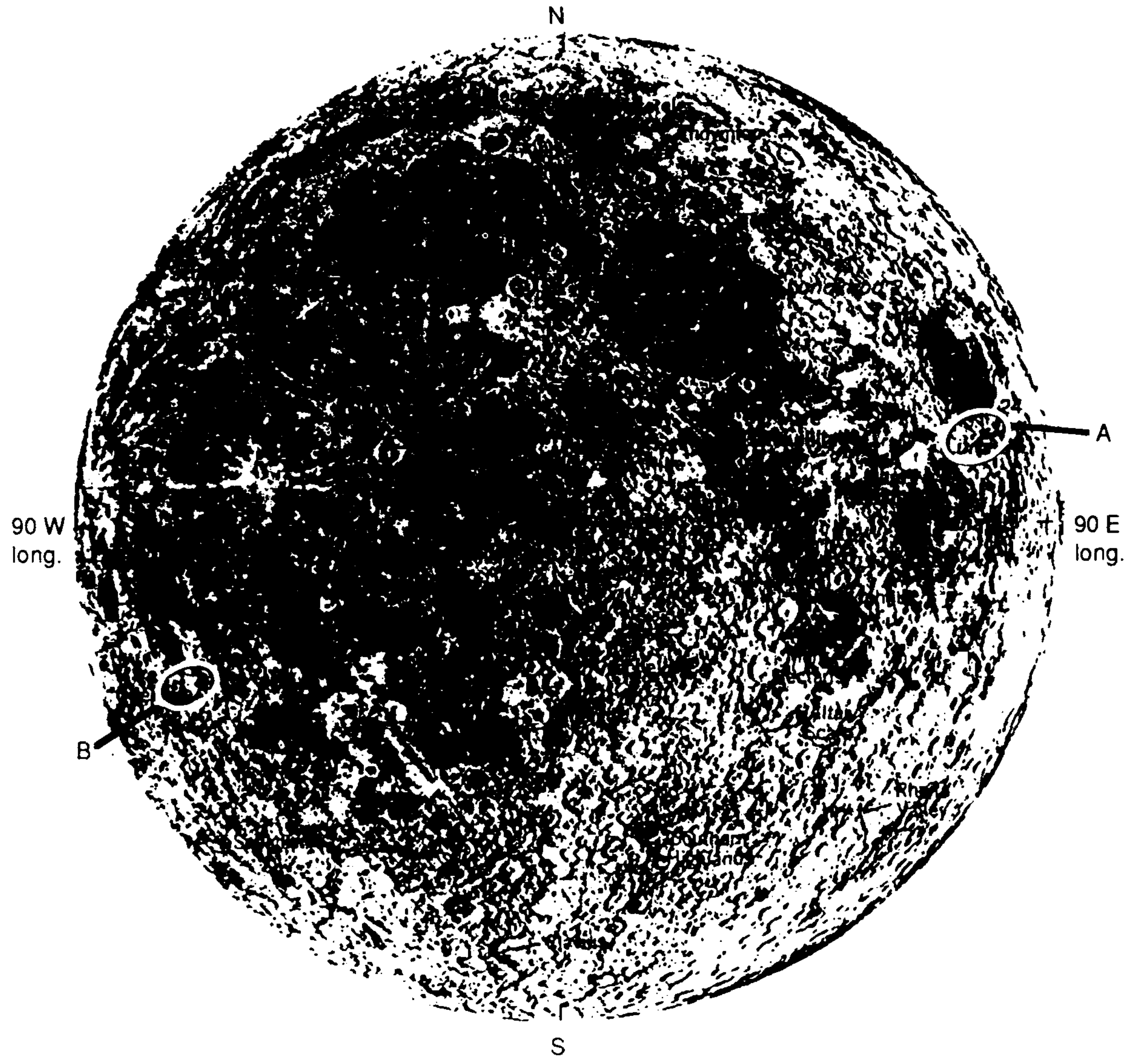
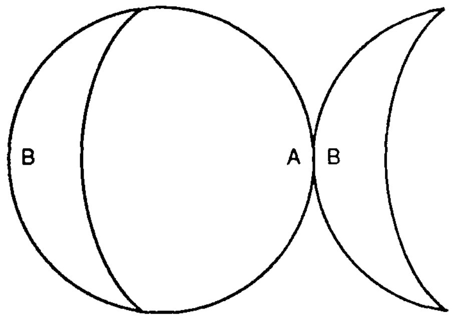



Figure 13. Map of near side of moon, showing the two regions used by Danjon.




Juxtaposition of two images

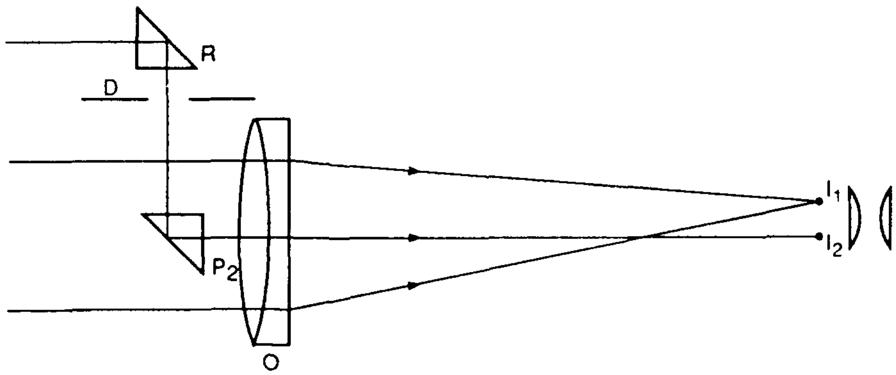



Figure 14. Schematic diagram of the cat's-eye photometer (from Reference [3]).

However, this ratio is also given in terms of the earth's reflectance properties as

$$\frac{I_E}{I_S} = \frac{1}{R_{EM}^2}\frac{d\sigma_E}{d\Omega}(\phi) = \frac{1}{R_{EM}^2}\,\pi R_E^2\,\frac{1}{\pi}p_E f_E(\phi)\;. \tag{6-3}$$

where $R_{EM}$ is the earth-moon distance, and $p_E$ and $f_E$ are the earth's geometric albedo and phase function. Upon equating (6-2) and (6-3) and solving for $p_E f_E$, we find

$$p_E f_E(\phi) = \left(\frac{R_{EM}}{R_E}\right)^2 D_{AB} f_M(\psi)\;. \tag{6-4}$$

Danjon used a separate series of comparisons between the intensities of the moon and the sun to determine $f_M$, as shown in Figure 15. Using this result, together with his observations of $D_{AB}$, he could determine $p_E f_E$ (or, equivalently, $I_E/I_S$) from Equation (6-2), as shown in Figure 16.

*Finally, the albedo is given as*

$$A = \int_{-\pi}^{\pi} p_E f_E(\phi)\sin(\phi)d\phi\;, \tag{6-5}$$

which is evaluated numerically after extrapolation to unmeasured phase angles. Separate values for $p_E$ and $q$ can be obtained by an extrapolation to $\phi = 0$.

Danjon's results show a number of interesting features. The daily means of the observations vary more widely than would be expected on the basis of the variation of measurements on a single night; this can be attributed to daily changes in cloud cover. A seasonal variation was also evident (see Figure 17), which is of the same shape as that of the ERBE measurements (Figure 1), but a factor of five larger in amplitude. Observations at several wavelengths also indicated that the earth's color changes with season. (These last two points were confirmed by the observations of Dubois [12].) No secular (annual or longer) variations were found.

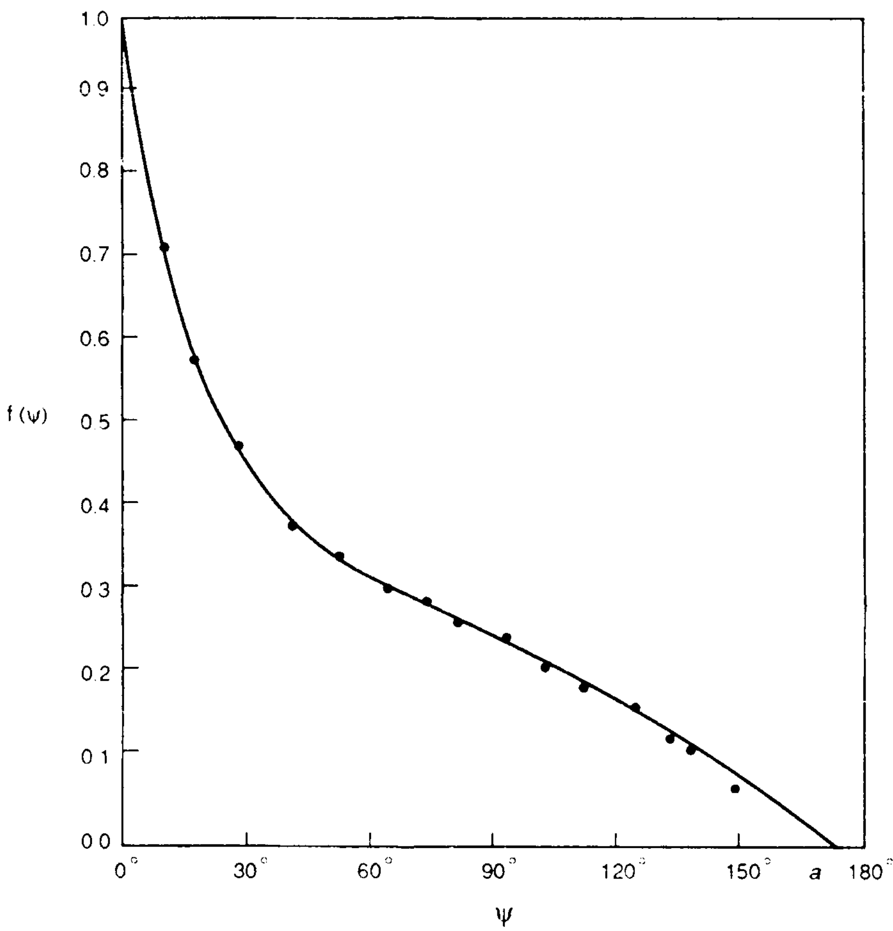



Figure 15. The phase function $f_M(\Psi)$ as determined by Danjon [3].

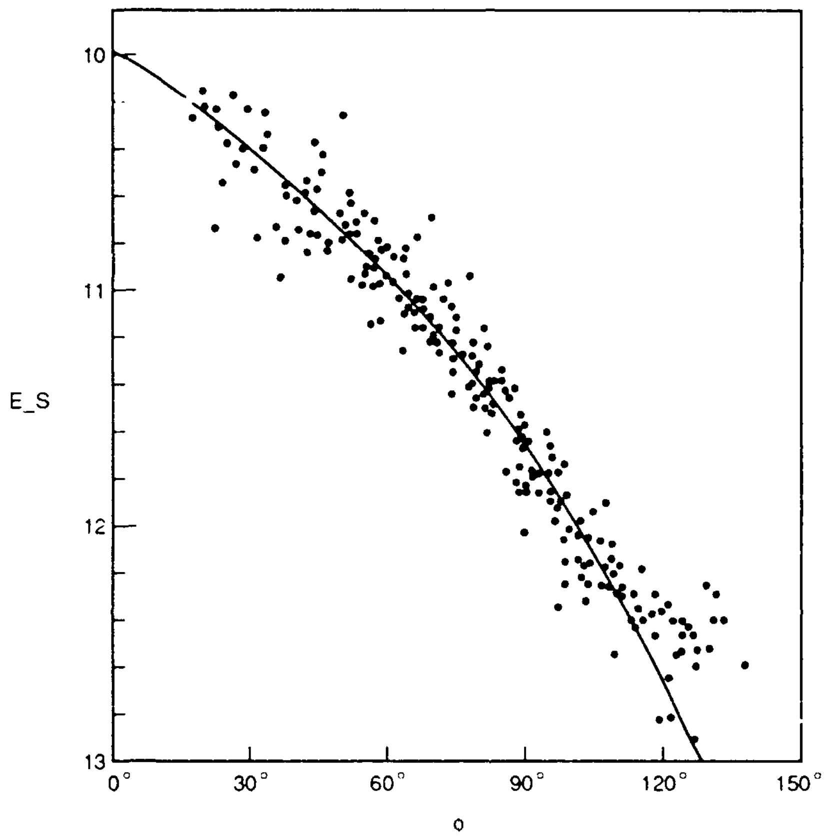



Figure 16. Abscissae, phase angle of the earth; ordinates, magnitude difference (as seen from the moon) between earth and sun. Corrected to mean distances and for seasonal variations (from Reference [3]).

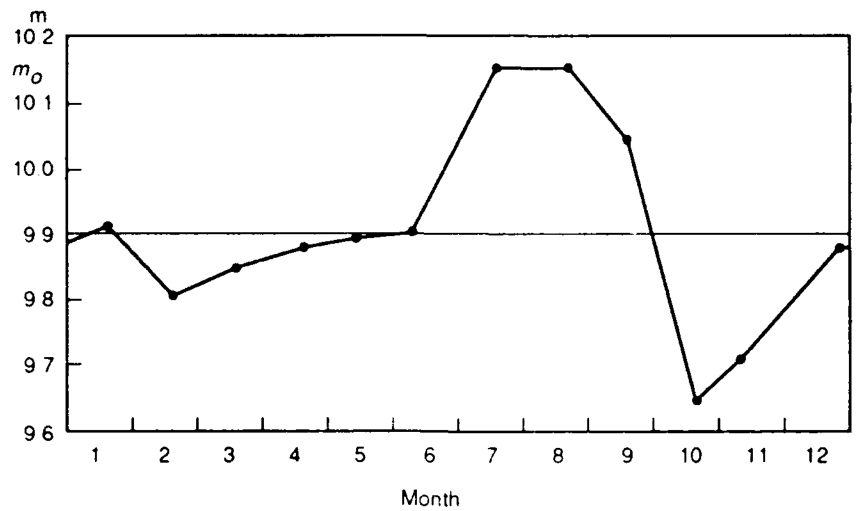



**Figure 17**. Variation of the monthly mean intensity of earthshine, expressed in magnitudes (from Reference [3])

In order to obtain the Bond albedo, Danjon's measurements in the visual must be corrected for the balance of the shortwave radiation (half of the sun's intensity is a wavelength greater than 0.7 $\mu$m). Estimates of this correction were made by Fritz [13], who also took into account that the Western Hemisphere most frequently observed by Danjon has a greater fraction of land than the globe as a whole. As the earth's albedo decreases with increasing wavelength (after all, the earth is sometimes called the "blue planet"), Fritz finds that Danjon's visual albedo of 0.40 corresponds to a Bond albedo of 0.36.

A second correction to the earthshine measurements must be made for the "opposition effect" present in the lunar reflectance properties; to our knowledge, this has not been considered previously. Observations of the moon [14, 15] (see Figure 18) show that the moon's phase function can rise by as much as a factor of 2 as $|\psi|$ decreases from 5 degrees to 0 (exact *backscattering). This enhancement* is caused by the porous nature of the lunar surface [11], and was unknown at Danjon's time since measurements close to $\psi = 0$ are hindered by lunar eclipses. (Note that the smallest angle measured in Danjon's lunar phase curve is only 11 degrees.) The extent of the small-angle rise varies over different regions of the lunar surface [14-15], but can easily be the 20 percent required to reduce Fritz's 0.36 to 0.29. This latter value is consistent with the ERBE satellite estimates of 0.30 [6].

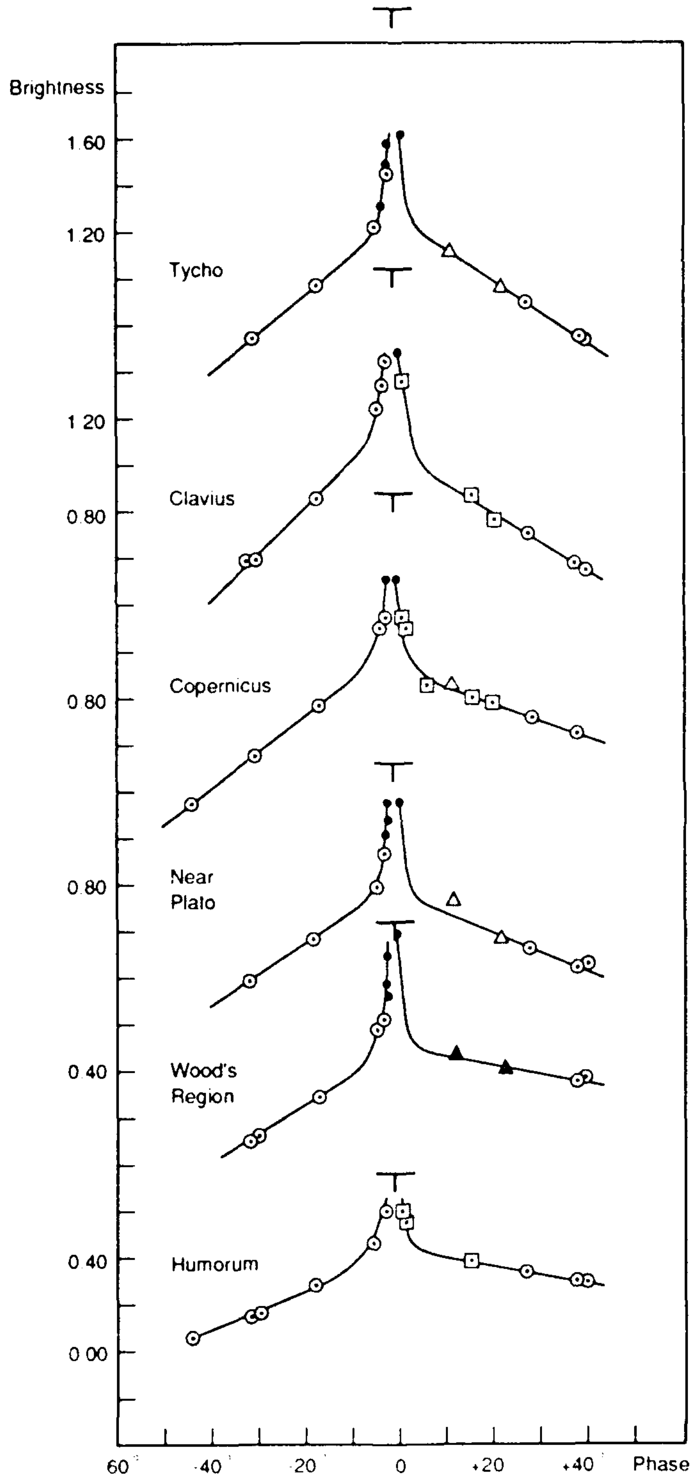



Figure 18. Brightness observed for various features on the moon as a function of phase. The T symbols include the extrapolation to zero phase (from Reference [13])

# 7 THE POTENTIAL FOR MODERN MEASUREMENTS

In the modern context, earthshine measurements have a number of attractive aspects. Ground-based measurements that integrate on a global scale are rare and the albedo is a basic parameter of the climate system. Earthshine measurements would complement more detailed satellite studies and could serve, at a minimum, as a cross check on the scene models used. That is, given meteorological data, these models make clear, non-trivial predictions of the time variation of the earthshine, which should be checked observationally.

Modern photometry can do significantly better than the state of the art in the 1920s. CCDs with 1024 × 1024 resolution and sufficient dynamic range to observe both the sunshine and the earthshine simultaneously can be purchased off the shelf. This would supplant Danjon's two-spot scheme. Together with a small (say 8") reflecting telescope, the total cost should be under $50,000. Two-dimensional imaging arrays that could extend observations to the near IR are also available, although at a somewhat greater cost. It might also be possible to illuminate the moon directly with a ground-based laser to measure the lunar geometric albedo. This latter is essential to deterring the absolute value of the earth's albedo but is unimportant if only changes are of interest.

Beyond a set of demonstration measurements, a long term monitoring program could make unique contributions to global change studies. Three observation sites spaced around the globe are sufficient for continuous coverage during the majority of the month when the earthshine is intense enough to be observed. Interannual variations of the albedo are clearly of great interest,

as are their correlations to global mean surface temperature. (As interannual temperature variations are $\leq 0.5$K, we might expect albedo variations $\lesssim 0.005$.) The phase function itself may be more sensitive to global warming than is the phase integral; this can, of course, be studied with models.

Today almost all GCMs give geographic models of surface albedos. Various parameterization schemes are used to capture changes in ice and snow cover, alteration of vegetation, etc. One of the results of running a GCM is the albedo determined as the difference between incoming solar radiation and the calculated outgoing infrared radiation. In this calculation the effects of assumptions about the surface and cloudiness are integrated over the globe. A measurement of the integrated albedo, such as that obtained from observations of earthshine, would provide a valuable check as to how well the model was representing the surface and cloud components of the albedo. A comparison of the seasonal variation in observed and computed albedo would provide another check on GCMs. If the observation could be maintained over a long interval, then the secular changes in observed and computed albedo could be compared.

Finally, there is the interesting possibility of an historical record of earthshine measurements spanning the 60 years since Danjon's work. Our preliminary investigations have turned up only the work of Dubois extending to 1958 [12]. However, if other data exist, they could provide a unique window on the secular change of the earth's climate. [We have recently learned of efforts at the University of Arizona (D. Huffman, private communication) aimed at duplicating Danjon's instrument and observing technique.]

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